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A SURVEY OF ATMOSPHERIC TURBULENCE CHARACTERISTICS

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19 August 1981

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This report reviews the literature on atmospheric turbulence in the planetary boundary layer, with the main emphasis on the lowest tens of meters. A table summarizes 26 references which discuss variations in intensity of turbulence. This intensity tends to be greater over rough surfaces than over smooth surfaces. Intensity of turbulence normally decreases when either altitude or atmospheric stability increases. High wind speeds are more likely to be associated with nearly neutral stability than with very stable or very

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unstable conditions. This report also includes a discussion of the autocorrelation function for the three components of the wind vector.

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I. INTRODUCTION

In studies of atmospheric turbulence, the positive x direction is usually defined as the direction of the mean horizontal wind, and the component of the wind in the x direction is the u component. The time average of u is denoted by $\bar{u}$, and $u' = u - \bar{u}$. The other horizontal coordinate is y, and the component of the wind in the y direction is v. By definition the time average of v, $\bar{v}$, is zero, and $v' = v$. The vertical coordinate is z, and the xyz system is a right-handed rectangular coordinate system. The vertical component of the wind is w. The quantities u' and v' are often referred to as the longitudinal and the lateral fluctuations of velocity, respectively. The corresponding standard deviations are σ_u and σ_v, where $\sigma_u = [\overline{(u')^2}]^{1/2}$ and $\sigma_v = [\overline{(v')^2}]^{1/2}$. The longitudinal intensity of turbulence is defined as $i_u = \sigma_u/\bar{u}$, and the lateral intensity is $i_v = \sigma_v/\bar{u}$.

In approximately the lowest 50m the horizontal wind stresses are nearly constant with height, and the wind direction is also nearly constant [Hess, 1959]. Then the following equation can be derived:

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{kz} \quad , \qquad (1)$$

where u_* is called the friction velocity and k is the von Karman constant. [Busch, 1973] discusses some different estimates of the von Karman constant, but it is now most commonly taken to be 0.4. [Panofsky, 1973] shows that the friction velocity at 100m is typically only about 10 per cent less than the surface value in mid-latitudes.

Integration of Equation (1) yields $u = (u_*/k)\ell n z$ plus a constant, but the logarithm of zero is undefined. Therefore, a roughness parameter, z_0, is introduced to obtain

$$\bar{u} = \frac{u_*}{k}\left(\ell n \frac{z}{z_o} - \psi\right) \quad , \qquad (2)$$

where ψ is a stability parameter. This wind law does not apply below $z = z_0$.

Tables of roughness parameter appear in [Hess, 1959] and [Frost et al., 1978]. Over ice z_0 may be less than 0.01cm. In high grass or wheat z_0 is typically a few centimeters. In a forest the roughness parameter is a fraction of a meter to one meter. In a city the roughness parameter is normally 1-4m.

ψ is zero under conditions of neutral stability and the simple logarithmic wind law, $\bar{u} = (u_*/k)\ \ell n\ (z/z_0)$ is obtained. Neutral stability exists when the lapse rate of temperature is equal to the adiabatic lapse rate. Under this condition a parcel of air displaced vertically experiences no buoyant acceleration.

When the temperature decreases with altitude more rapidly than the adiabatic lapse rate, the atmosphere is unstable and ψ is positive. [Blackadar et al, 1974] contains information for estimating ψ, which may be greater than 1.7 for a very unstable atmosphere.

In a stable atmosphere ψ is negative and can be expected to have a magnitude less than 1.0.

[Pasquill, 1961] devised a classification scheme for estimating atmospheric stability from surface (10m) wind speeds and amount of heating or cooling. Pasquill stability classes range from A through F, where A is very unstable, D is neutral, and F is the most stable class. According to Pasquill's classification, as wind speeds increase the atmosphere approaches neutral stability. When speeds are greater than 6m/sec, conditions are neutral unless there is strong insolation, in which case the atmosphere may be slightly unstable. At night, cloud cover less than 50 percent permits enough cooling of the surface that the atmosphere is in the most stable F Pasquill category for wind speeds of 3m/sec or less. Whenever there is a heavy overcast of clouds, the atmosphere is neutral regardless of wind speed or time of day or night. During the day the amount of insolation depends upon sun angle and cloud amount and type. [Luna and Church, 1972] outline details of a procedure for determining insolation from standard meteorological observations.

This report contains an extensive discussion of variations of intensity of turbulence with atmospheric conditions, surface roughness, and altitude. There is also a discussion of the autocorrelation function. Finally, other conditions affecting turbulence are discussed.

II. INTENSITY OF TURBULENCE

This section presents a discussion of some measurements of intensity of turbulence made at Redstone Arsenal, Alabama, and a survey of the literature describing results obtained by others.

Table I contains $\bar{u}$ and the longitudinal and lateral intensities of turbulence measured in August 1973 under unstable atmospheric conditions. The site was the Army Gas Dynamics Laser Range. It consisted of a grass-covered plot, 61m wide and 655m long, with trees lining each side. Five towers were located 137m apart, and a sixth was 9.1m from the middle tower on a line perpendicular to the line of the other five towers. More information can be found in [Stewart, 1975].

Both longitudinal and lateral intensities of turbulence vary with height. At 10m, i_u varies from 0.25 to 0.98 and has a mean of 0.56, while i_v has a much larger variation from 0.16 to 1.41 and a mean of 0.53. If a mean is taken of the individual ratios i_v/i_u, 0.94 is obtained at 10m. At 6m i_v/i_u has a mean of 1.04. The mean i_v of 0.61 is also greater than the mean i_u, which is 0.59. At 2m the mean of the ratios i_v/i_u is 1.10. At 2m i_v varies from 0.21 to 3.33 and has a mean of 0.85. At 2m i_u also has a wide variation from 0.27 to 2.09, and the mean is 0.77. Note that $i_v/i_u = \sigma_v/\sigma_u$. It follows that under the conditions of these measurements the standard deviation of the lateral component of the wind is the same order of magnitude as the standard deviation of the longitudinal component.

TABLE I. INTENSITIES OF TURBULENCE AT TEST AREA 5 ON REDSTONE ARSENAL ($\bar{u}$ = m/sec)

Case		Tower 1			Tower 2			Tower 3W			Tower 3E			Tower 4			Tower 5		
		10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m
I	$\bar{u}$	1.46	1.28	0.91	1.34	1.29	1.15	1.64	1.66	1.46	1.76	1.61	1.17	1.62	1.58	1.41	1.32	1.23	1.16
	i_u	0.50	0.57	0.74	0.63	0.72	0.85	0.58	0.64	0.71	0.58	0.60	0.79	0.60	0.66	0.79	0.74	0.81	0.78
	i_v	0.16	0.38	0.58	0.32	0.35	0.40	0.58	0.54	0.62	0.57	0.60	0.81	0.43	0.40	0.42	0.66	0.60	0.57
II	$\bar{u}$	1.05	1.08	0.92	1.45	1.43	1.16	1.13	0.97	0.76	1.02	0.83	0.21	0.78	0.61	0.45	0.87	0.70	0.70
	i_u	0.58	0.64	0.63	0.41	0.45	0.47	0.48	0.54	0.67	0.49	0.51	2.00	0.85	0.75	0.89	0.57	0.73	0.90
	i_v	0.45	0.48	0.81	0.32	0.38	0.46	0.52	0.80	1.06	0.77	0.93	2.62	0.67	0.98	1.51	0.67	0.83	0.76
III	$\bar{u}$	1.77	2.03	1.66	1.34	1.42	1.14	1.51	1.20	0.90	1.50	1.16	0.49	1.62	1.14	1.15	1.85	1.51	1.45
	i_u	0.41	0.47	0.54	0.63	0.58	0.61	0.56	0.71	0.88	0.63	0.69	1.61	0.58	0.66	0.61	0.69	0.64	0.64
	i_v	0.52	0.44	0.55	0.54	0.49	0.56	0.38	0.66	0.92	0.46	0.71	1.40	0.38	0.56	0.65	0.39	0.45	0.46
IV	$\bar{u}$	2.13	2.02	1.78	1.41	1.30	1.07	1.92	2.25	1.99	2.04	2.19	1.63	1.69	1.72	1.56	1.78	1.75	1.48
	i_u	0.47	0.50	0.52	0.65	0.69	0.78	0.54	0.46	0.50	0.53	0.50	0.58	0.41	0.47	0.52	0.39	0.39	0.45
	i_v	0.52	0.68	0.80	0.78	0.95	1.13	0.36	0.29	0.38	0.35	0.32	0.35	0.39	0.34	0.42	0.49	0.43	0.61
V	$\bar{u}$	2.92	2.80	2.52	2.16	2.09	2.01	2.04	2.33	2.14	2.20	2.30	1.81	1.69	1.72	1.47	1.31	1.39	1.11
	i_u	0.25	0.24	0.27	0.33	0.32	0.31	0.34	0.35	0.36	0.32	0.32	0.40	0.42	0.45	0.48	0.46	0.50	0.59
	i_v	0.16	0.17	0.21	0.26	0.32	0.35	0.26	0.21	0.24	0.21	0.20	0.22	0.36	0.25	0.33	0.68	0.64	0.93

TABLE I. (Concluded)

Case		Tower 1			Tower 2			Tower 3W			Tower 3E			Tower 4			Tower 5		
		10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m	10m	6m	2m
VI	$\bar{u}$	1.45	1.43	1.25	1.95	1.72	1.38	1.02	0.90	0.59	1.04	0.84	0.45	1.44	1.13	0.90	1.51	1.30	1.16
	i_u	0.48	0.51	0.53	0.40	0.41	0.41	0.73	0.64	0.93	0.67	0.69	1.20	0.53	0.57	0.58	0.51	0.55	0.58
	i_v	0.62	0.66	0.70	0.47	0.57	0.74	0.60	0.90	1.35	0.68	0.84	1.49	0.47	0.58	0.69	0.52	0.49	0.58
VII	$\bar{u}$	2.24	2.42	2.11	1.99	1.85	1.51	1.67	1.44	1.03	1.69	1.36	0.57	1.59	1.25	0.98	1.67	1.48	1.25
	i_u	0.38	0.40	0.51	0.69	0.66	0.73	0.72	0.70	0.86	0.67	0.74	1.70	0.50	0.50	0.82	0.62	0.63	0.67
	i_v	0.41	0.42	0.60	0.32	0.35	0.48	0.35	0.55	0.78	0.50	0.64	1.17	0.40	0.60	1.00	0.72	0.80	0.80
VIII	$\bar{u}$	1.60	1.57	1.48	1.13	1.08	0.87	1.16	1.10	0.87	1.28	1.07	0.68	1.55	1.46	1.19	1.30	1.20	0.97
	i_u	0.47	0.54	0.55	0.86	0.86	1.13	0.66	0.70	0.87	0.65	0.71	1.10	0.44	0.47	0.54	0.57	0.66	0.86
	i_v	0.58	0.73	0.95	0.68	0.71	0.93	0.71	0.88	0.99	0.68	0.89	0.91	0.41	0.40	0.46	0.53	0.70	0.93
IX	$\bar{u}$	1.76	1.87	1.72	2.05	1.81	1.55	0.99	0.96	0.57	1.00	0.91	0.33	1.40	1.15	0.98	1.46	1.30	1.23
	i_u	0.44	0.56	0.59	0.54	0.54	0.55	0.98	0.93	1.47	0.88	0.88	2.09	0.71	0.72	0.92	0.58	0.71	0.78
	i_v	0.70	0.73	0.80	0.40	0.53	0.63	1.41	1.60	2.57	1.33	1.46	3.33	0.84	1.06	1.26	0.55	0.70	0.73

There is no agreement among diverse sets of observations on the magnitude of σ_v/σ_u. This probably represents real changes in atmospheric conditions. For example, [Swanson and Cramer's, 1965] Table 6 contains mean values of σ_v/σ_u at 1m, and the magnitude of this ratio varies from 0.44 to 1.03. [Mayer's, 1981] Table 3 summarizes 4 sets of measurements made in a spruce forest, and here the ratio σ_v/σ_u varies from 0.65 to 0.86.

There is evidence that the ratio σ_v/σ_u depends upon stability. For example, [Wyngaard and Clifford, 1977] include in their Table 1 mean values of $(\sigma_u/\overline{u})^2$ and $(\sigma_v/\overline{u})^2$ for 5.7m over a flat uniform Kansas plain. For very unstable conditions the mean σ_v divided by the mean σ_u is 1.32. For moderately unstable conditions this ratio is 0.95, but when the atmosphere is stable the ratio is 0.74. From Table 2 of [Panofsky et al., 1978] one obtains magnitudes of the ratio σ_v/σ_u of 0.72 to 1.18 for unstable conditions at a height of 2m. [Champagne et al., 1977] give σ_v and σ_u for 4m above flat farm land during unstable conditions. For all of the 4 sets of data in their Table 1 the ratio σ_v/σ_u is greater than 1.0 and in one test run σ_v was more than 50 percent greater then σ_u. On the other hand, [Ariel and Nadezhina, 1976] summarized field measurements for neutral stratification, and σ_v/σ_u varied from 0.74 to 0.89. [Skibin, 1972] obtained σ_v/σ_u of 0.48 in an experiment where stability was considered neutral according to its Pasquill category, and σ_v/σ_u was 0.26 for slightly unstable conditions. For more unstable conditions Skibin obtained σ_v/σ_u from 0.27 to 0.37. The problem of making these data consistent with other investigations may be the inexactness of the Pasquill method [Luna and Church, 1972].

The relative magnitudes of σ_v and σ_u also depend upon height. [Frost et al., 1978] consider a neutrally stable atmosphere, and from their equation 4.26, which applies at a height of 10m, $\sigma_v/\sigma_u = 0.64$ is obtained. They then proceed to give a typical example where this ratio increases to 1.00 at 600m, above which $\sigma_v = \sigma_u$. According to [Dickson and Angell's, 1968] Figure 5 $\sigma_v = \sigma_u$ at 2km, but σ_v is from 0.7 σ_u to 0.9 σ_u at 0.5km. [Bowne and Ball, 1970] list means of σ_v/u_* and σ_u/u_* taken from different stabilities for two levels in a rural location and two levels in an urban location. Outside the city the mean σ_v divided by the mean σ_u is 0.76 at 12.2m and 0.97 at 61m. Inside the city the ratio is 0.76 at 15.3m and 0.60 at 53.3m. [Bradley's, 1980] Table 3 describes measurements at the crest of a 170-m hill; these data do not show a consistent change with height. σ_v/σ_u ranges from 0.65 to 0.74 at 9m and from 0.77 to 0.82 at 16m. At 25m the ratios vary from 1.09 to 1.17; however, at 87m the magnitudes of σ_v/σ_u are lower than at 25m and range from 0.71 to 1.02.

Table II summarizes some of the literature describing measurements of i_u and σ_u/u_*. The magnitude of i_u varies from less than 0.1 to more than 1.0, but the most typical values are between 0.1 and 0.4. The values of σ_u/u_* are commonly between 1.5 and 3.5.

The intensity of turbulence normally decreases as stability increases. This is illustrated in the work of [Swanson and Cramer, 1965] who made measurements at White Sands Missile Range. Their tower stood on a smooth plot surrounded by ground containing small, uniformly distributed sand dunes. They measured temperature, wind speed, and wind direction at nine levels from 4.6 to 62.0m. Only observation periods with regular temperature profiles and

TABLE II. SUMMARY OF LONGITUDINAL INTENSITIES OF TURBULENCE

Reference	Surface	Stability	Height	$\sigma_u / \overline{u}$	σ_u/u_*
Cramer (1959)	Prairie Grass with $z_o < 1$ cm		2m	0.16–0.34	
Swanson and Cramer (1965)	Smooth ground surrounded by sand dunes	Unstable	4.6m 33.9m 62.0m	0.22–0.42 0.18–0.33 0.17–0.30	
		Neutral	4.6m 33.9m 62.0m	0.21–0.27 0.15–0.18 0.13–0.16	
		Stable	4.6m 33.9m 62.0m	0.19–0.25 0.11–0.13 0.10–0.13	
Cramer (1967)	grass		16m 40m		2.5 2.0
Dickson and Angell (1968)		Unstable	2000m 500m	0.22 0.43	
Allen (1968)	Japanese Larch Plantation		1.15m 10.4 m	0.51 0.47	
Arya and Plate (1969)		Stable	2m–4m		1.8–3.2
Fichtl and McVehil (1970)		Neutral			$2.227(z/18)^{-0.315}$
		Unstable			$1.897(z/18)^{-0.07}$ (z in meters)

TABLE II. (Continued)

Reference	Surface	Stability	Height	$\sigma_u / \overline{u}$	σ_u/u_*
Bowne and Ball (1970)	Rural Urban		12.2m 61.0m 15.3m 53.3m	0.01-0.58 0.08-1.29 0.29-1.34 0.15-0.52	1.38-3.19 1.82-3.54 2.00-8.01 1.74-3.69
Grimm (1971)	z_o = 1.1cm z_o = 2.6cm	neutral-slightly stable unstable			2.36 3.31
Cionco (1972)	in and above canopies		within canopies just above canopies	0.32-0.84 0.28-0.47	
Skibin (1972)		unstable neutral		.309-.852 .356	
McBean and MacPherson (1976)	Over and near Lake Ontario		30-300m	0.083	
Ariel and Nadezhina (1976)		neutral			1.7-3.2
Panofsky et al.(1977)	flat land	unstable			2.8-5.5
Rayment and Caughey (1977)	typical rural	unstable	91m	0.23	

TABLE II. (Continued)

Reference	Surface	Stability	Height	$\sigma_u / \overline{u}$	σ_u/u_*
Wyngaard and Clifford (1977)	flat, uniform Kansas plain	very unstable moderately unstable moderately stable	5.7m	0.22 0.20 0.16	
Champagne et al. (1977)	flat farm land	unstable	4m	0.18	
Panofsky et al. (1978)	flat, uniform	unstable	2m		2.1-4.2
Bradley and Antonia (1979)		z/L <-0.1 (unstable)	5m		3.5-9.5
Binkowski (1979)		unstable neutral stable			2.0-4.8 2.0-3.8 1.8-2.6
SethuRaman and Raynor (1980)	ocean beach		8m 8m	0.02-0.19 0.04-0.38	
Bradley (1980)	top of 170m hill	neutral	87m 9m	0.10 0.38	
Mayer (1981)	spruce forest		at tree top	0.59-1.30	1.8-2.1

TABLE II. (Concluded)

Reference	Surface	Stability	Height	$\sigma_u/\overline{u}$	σ_u/u_*
Hanna (1981)		unstable	100-900m	0.084-1.0	
Stewart (unpublished)	grass covered cut through forest	unstable	10m 6m 2m	0.25-0.98 0.24-0.93 0.27-2.09	
Webster and Burling (1981)	15cm grass	neutral	2m	0.13-0.16	

4.6-m wind speeds of at least 4 mi/hr(1.8m/sec) were included in the data base. They used the rather unusual definition that conditions were neutral if the magnitude of the temperature difference between 4.6m and 62.0m was not greater than 0.56°C. On the basis of 824 observation periods Swanson and Cramer found that when the atmosphere was less stable than neutral, the intensity of turbulence was greater than when the atmosphere was stable. Intensities were intermediate during neutral stability at each of the nine levels. For example, at 62.0m for wind speeds from 1.8 to 3.1 m/sec, i_u = 0.30 for unstable conditions, i_u = 0.16 for neutral conditions, and i_u = 0.13 for stable conditions. At 4.6m the corresponding intensities are 0.42, 0.27, and 0.25, respectively. At higher wind speeds turbulence intensities are smaller. For wind speeds greater than 4.46m/sec under unstable conditions the longitudinal intensities of turbulence are 0.22 and 0.17 at 4.6 and 62.0m, respectively. Under stable conditions the corresponding magnitudes of i_u are 0.19 and 0.10.

[Skibin, 1972] includes a table of five experimental sets of observations made in connection with an atmospheric dispersion study. One of these was under Pasquill stability category D, or neutral, and had a value of 0.356 for $\sigma_u/\overline{u}$. In a slightly unstable case $\sigma_u/\overline{u}$ was 0.414. For more unstable cases, longitudinal turbulence intensities were 0.309, 0.403, and 0.852.

[Wyngaard and Clifford, 1977] summarize earlier work in their Table 1 in which they list the mean values of $\sigma_u^2/\overline{u}^2$ over a flat, uniform Kansas plain. Their data show that for very unstable conditions $\sigma_u/\overline{u}$ is 0.22 and is 0.20 for moderately unstable conditions. In a moderately stable atmosphere $\sigma_u/\overline{u}$ is only 0.16.

In [Binkowski's, 1979] Figure 4 many cases of σ_u/u_* are plotted graphically as a function of stability. Near neutral conditions the values of σ_u/u_* are mostly near 2.5, and under stable conditions the mean is nearer 2.2. For an unstable atmosphere the mean σ_u/u_* becomes larger as the atmosphere becomes less stable, and is near 4.0 for a very unstable atmosphere.

[Grimm, 1971] examined 15 cases which were either neutral or slightly stable, where z_o = 1.1cm, and obtained a mean σ_u/u_* of 2.36 for heights from 8 to 32m. In 44 unstable cases, where z_o = 2.6cm, the mean σ_u/u_* was 3.31.

The intensity of turbulence depends upon the roughness of the underlying surface, as well as upon the stability. Usually the intensity increases as stability decreases or as the roughness of the underlying surface increases. [Hanna, 1981] points out that an exception to this rule may occur in very stable, light wind conditions.

[Bowne and Ball, 1970] obtained observations of turbulent wind fluctuations on a tower in downtown Fort Wayne, Indiana, and on another tower in a nearby rural setting. The roughness was larger in the city, and the urban heat island reduced atmospheric stability, especially at lower levels. Turbulence was more intense in the rougher and less stable urban environment. The lower levels on the urban and rural towers were 15.3 and 12.2m, respectively. In 16 of 19 tests $\sigma_u/\overline{u}$ at the lower level was greater on the urban than on the rural tower, and in many of these 16 tests the difference

was quite large. The upper levels on the urban and rural towers were 53.3 and 61.0m, respectively. The longitudinal intensity of turbulence in the urban location was greater in 14 of 18 tests. On the other hand when one considers σ_u/u_* one finds that at the upper level the mean of 2.48 in the urban location is only slightly larger than the 2.42 for the rural setting. At the lower level in the city the mean σ_u/u_* of 4.16 was much larger than the 2.47 measured outside the city.

[SethuRaman and Raynor, 1980] compared longitudinal intensities of turbulence at a height of 8m over the Atlantic Ocean, 5km from Long Island, New York, with simultaneous measurements at 8m above the beach. As with previously discussed studies over land, intensity of turbulence over the ocean decreases as stability increases. An overall average $\sigma_u/\bar{u}$ over the ocean is near 0.09, with a variation from approximately 0.02 to 0.19. The behavior of the ratio of longitudinal intensity of turbulence over the ocean to that over the beach depends upon whether the flow is basically onshore, offshore, or along the shore. When flow is offshore the intensity of turbulence over the ocean is approximately the same as that over land. When flow is along the shore, turbulence intensity over the ocean is about half that over land when the land wind speed is greater than 6m/sec; but for wind speeds less than 3m/sec turbulence intensity over the ocean is greater than that over land. For onshore winds, the intensity of turbulence over land and water is about the same when the wind speed over land is greater than 12m/sec, and there is a minimum ratio of oceanic-to-land turbulence of 0.5 near 10m/sec. Below 10m/sec the ratio of intensity over the ocean to that over land increases to almost 2 as wind speed decreases for onshore flow.

[Allen's 1968] Table 4 summarizes measurements made near Itchaca, New York, in a plantation of Japanese larch which had a mean height of 10.40m. At this height intensity of turbulence was 0.47, and it increased to 0.57 at 7.25m. Variation of intensity was irregular down to 1.15m where it was 0.51.

[Dionco, 1972] discusses canopies such as rice paddies, wheat fields, and forests. Just above the different types of canopies the intensity of turbulence varies from 0.28 to 0.47, but within canopies intensities range from 0.32 to 0.84.

[Mayer, 1981] found that the longitudinal intensity of turbulence at the top of a spruce forest varied from 0.59 to 1.30.

In the previous discussions of variation of intensity of turbulence with surface roughness and with atmospheric stability, the reader may have noticed that there also appeared to be variations with altitude. In general, $\sigma_u/\bar{u}$ can be expected to decrease as altitude increases. There is also evidence that σ_u decreases with altitude.

[Fichtl and McVehil, 1970] considered a large amount of data to develop their Table 2 in which the ratio $\sigma_u/(B_u^{1/2} u_{*o})$ is equal to 2.227 under neutral conditions and 1.897 under unstable conditions, where u_{*o} is the surface friction velocity. Fichtl and McVehil's Table 1 gives B_u as a function

of z in meters. For neutral conditions, $B_u = (z/18)^{-0.63}$, and for unstable conditions, $B_u = (z/18)^{-0.14}$. Thus, one obtains $\sigma_u/u_{*0} = 2.227 (z/18)^{-0.315}$ for neutral conditions and $\sigma_u/u_{*0} = 1.897(z/18)^{-0.07}$ under unstable conditions.

[Swanson and Cramer, 1965] analyzed both the longitudinal and lateral intensities of turbulence on a 62-m meteorological tower over a two-year period at White Sands Missile Range. Both intensities decreased with height in all thermal stratifications. Swanson and Cramer found that the decrease could be expressed as z to a power which varied from -0.1 to -0.3. The magnitude of the exponent is larger for more stable conditions. They also found that turbulent intensities at all heights and in all thermal stratifications tended to be inversely proportional to the mean wind speed.

[DeLarrinaga, 1972] tested the power law proposed by [Swanson and Cramer, 1965] on wind measurements from two urban sites in Liverpool. A captive balloon was used, and the upper anemometer was at 305m. The height of a lower anemometer was varied. These urban measurements verified that $\sigma_u/\bar{u}$ decreased as z increased. DeLarrinaga fitted the data to the power law described by Swanson and Cramer and obtained exponents from -0.14 to -0.36.

[Bowne and Ball, 1970] made simultaneous measurements at a rural and urban site. At the rural site wind measurements were at 12.2m and 61.0m. At the urban site instruments were at 15.3m and 53.3m. At the urban locations there were 18 sets of data where both upper and lower values of $\sigma_u/\bar{u}$ were available; in all 18 cases the intensity of the lower level turbulence was greater than the intensity of the upper level turbulence. In 14 of 17 sets of measurements from the rural site the intensity of turbulence was greater at the lower level than at the upper level.

[Petit et al., 1976] show a plot of $\sigma_u/\bar{u}$ within and above a forest in their Figure 5. Tnere is some irregularity within the forest to approximately 2m above tree top and then a decrease with altitude.

[Bradley, 1980] made wind measurements on a tower placed on top of a 170-m hill during atmospheric conditions associated with neutral stability. In the three sample cases in Bradley's Figure 3, $\sigma_u/\bar{u}$ at 9m is approximtely 3 times $\sigma_u/\bar{u}$ at 87m. In the text they define an intensity of turbulence as $[(1/3)\ (\overline{u'^2} + \overline{v'^2} + \overline{w'^2})]^{\frac{1}{2}}\ /\bar{u}$ and consider means of this intensity for all data: 0.326 for 9m; 0.168 for 25m; and 0.120 for 87m.

[Dickson and Angell, 1968] contain figures of σ_u and $\sigma_u/\bar{u}$ as functions of height to 2km, and both quantities decrease with increasing altitude during the one summer of data plotted; however, σ_u only decreases slightly, and σ_v increases slightly with increasing altitude. On the other hand, $\sigma_u/\bar{u}$ decreases from 0.43 at 0.5km to 0.22 at 2km and $\sigma_v/\bar{u}$ decreases from 0.33 to 0.21.

[Duchéne-Marullaz, 1975] found that longitudinal intensities of turbulence in a suburban area near Nantes, France, decreased from 0.30 at 10m to 0.20 at 60m for SSW and SW winds. The decrease was from 0.28 at 10m to 0.17 at 60m for westerly winds.

III. AUTOCORRELATIONS

In this section, a detailed discussion of autocorrelation functions is prefaced by an explanation of their importance. Equations are derived for the variance of the difference between $u(t+\tau)$ and $u(t)$ and for the variance of a random component of turbulence which is uncorrelated with the turbulent fluctuation either at that time or at another time. Each of these equations contains an autocorrelation function $R(\tau)$ which is the correlation between the values of u at two times separated by the time interval τ.

Let Δu be defined by the equation

$$\Delta u = u(t+\tau) - u(t) \tag{3a}$$

where t is time. If $\overline{u}$ is assumed to be constant throughout the time period being studied, Equation (3a) can be rewritten as

$$\Delta u = u'(t+\tau) - u'(t). \tag{3b}$$

Both sides of Equation (3b) may be squared and averaged to obtain

$$\overline{(\Delta u)^2} = \overline{[u'(t+\tau)]^2} + \overline{[u'(t)]^2} - 2\,\overline{u'(t+\tau)u'(t)}. \tag{4a}$$

Each of the first two terms on the right hand side of Equation (4a) is equal to σ_u^2, and the third term equals $2\sigma_u^2(R(\tau))$. Thus is obtained

$$\overline{(\Delta u)^2} = 2\,\sigma_u^2(1-R(\tau)). \tag{4b}$$

When $\tau = 0$, $R(\tau) = 1$, and $\overline{(\Delta u)^2}$ is zero as expected. When τ is very large, the autocorrelation is zero, and the variance of Δu is twice the variance of u.

It is sometimes convenient to assume a linear relationship between $u'(t+\tau)$ and $u'(t)$ and to write

$$u'(t+\tau) = u'(t)R(\tau) + u''(t) \tag{5a}$$

where u'' is independent of u'. [Hanna, 1979] recommends such a relationship when the coordinate system is following an air parcel, but it can also be used to describe behavior at a point which is stationary relative to the earth. One can rewrite Equation (5a) as

$$u''(t) = u'(t+\tau) - u'(t)R(\tau). \tag{5b}$$

If squares and averages are made of both sides of Equation (5b) the following is obtained:

$$\sigma_{u''}^2 = \sigma_{u'}^2 R^2(\tau) + \sigma_{u'}^2 - 2R(\tau)\overline{u'(t)u'(t+\tau)}. \quad (6a)$$

Since $\overline{u'(t)u'(t+\tau)}$ is equal to $R(\tau)\sigma_{u'}^2$, Equation (6a) can be rewritten as

$$\sigma_{u''}^2 = \sigma_{u'}^2(1-R^2(\tau)). \quad (6b)$$

It is obvious that for very large values of τ the variance of u" is equal to the variance of u'.

Information can be obtained about spatial correlations by applying Taylor's hypothesis which relates spatial correlations to time lag correlations. This frozen field or frozen turbulence approximation makes the substitution $\tau = \Delta x/\overline{U}$, where Δx is the spatial lag in the direction of $\overline{U}$. It is assumed that the turbulence is homogeneous in the x direction and stationary in time [Lumley and Panofsky, 1964] or [Webster and Burling, 1981].

In order to test the relationship between space and time correlation functions, [Cramer, 1959] carefully selected six experiments from Project Prairie Grass. In each one, the observed wind direction was within 25 degrees of the longitudinal axis of the instrument array. In Cramer's Figure 11 for daytime and Figure 12 for nighttime experiments, both spatial and temporal correlations of u are plotted. The abscissa is τ for the autocorrelation data and $\Delta x/\overline{U}$ for the spatial data. The ordinate is (1-R). Cramer's data show close agreement between spatial correlations and temporal autocorrelations during both day and night for the 60 seconds of data which were plotted. This shows that Taylor's hypothesis is useful in the atmosphere as well as in the laboratory.

Figures 5.20 and 5.21 of [Lumley and Panofsky, 1964] depict the autocorrelation functions of both u and v lagged in both space and time for a time period of 20 seconds. For the daytime observations agreement is extremely close, but time autocorrelations tend to be slightly larger than space autocorrelations. During a typical night period space and time autocorrelations of v are also close, but time autocorrelations of u are consistently much larger than space autocorrelations.

[Elderkin and Powell, 1971] also tested the space-time relationships in atmospheric turbulence. Their Table 1 contains $R(\tau)$ and $R(\Delta x)$ for u', v', and w'. According to Taylor's hypothesis, $R(\tau) = R(\Delta x)$ if $\tau = \Delta x/\overline{U}$. Their table goes to 252m, which corresponds to almost 40 seconds for the applicable $\overline{U}$ of 6.4m/sec. For u' the correlations $R(\Delta x)$ and $R(\tau)$ are nearly identical to 48m (7.5 sec), but beyond this point some of the R's differ by 20 percent or more. For v' and w' the R's diverge considerably after 4 sec. Beyond a few seconds all three spatial correlations are higher than the corresponding time correlations. Therefore, one must use some caution when applying Taylor's widely used hypothesis.

[Tennekes and Lumley, 1972], in their Figure 8.2, illustrate idealized behavior of R on a spatial scale. The spatial correlation of the u component decreases smoothly from unity at zero separation to zero at large separations. The autocorrelation of the v component decreases smoothly to zero and becomes slightly negative before leveling off at zero for large distances.

Actual autocorrelations may have quite irregular variations with time. This is illustrated in Figures 1 and 2 of [Stewart, 1975]. The figures contain a representative sample of autocorrelations which were computed for 1-sec intervals from lag zero through lag 120 for u, v, and w at 10m. In some sets of measurements analyzed by Stewart, the autocorrelation function crossed zero several times. On the other hand, in some instances the autocorrelations of u and v did not reach zero during the entire 120 sec for which the computations were done. The autocorrelation of the w component usually reached zero in 20 seconds or less.

[Mackey and Ko, 1975] show the autocorrelation functions up to 150 sec for the longitudinal fluctuations at 13, 28, 43, and 61m in a typhoon. These curves decrease rapidly for approximately 25 sec and then fluctuate irregularly about a mean value. The 61-m autocorrelation levels off to fluctuate about a mean near 0.3, and the 13-m autocorrelation levels off to fluctuate about 0.15. The 28-m and 48-m autocorrelations fall between the ones for 61m and 13m.

In spite of the irregularities in many autocorrelation functions which are obtained from measurements, many investigators have attempted to develop simple analytical approximations. One of the simplest and most widely used approximations for the u component is

$$R(\tau) = \exp(-\tau/T) \tag{7}$$

where

$$T = \int_0^\infty R(\tau)d\tau \tag{8}$$

is an integral time scale. [Hanna, 1979] averaged $R(\tau)$ for the u component over several unstable runs made above flat farm land in Minnesota. This autocorrelation function was plotted versus τ for 60 sec and compared with the curve obtained by fitting observations to Equation (7). This approximation appeared good to within 20 percent for the averaged unstable runs.

[Fichtl and McVehil, 1970] considered an exponential equation similar to Equation (7) and applied it to space-lagged autocorrelations. They examined a large amount of data from unstable and neutral atmospheres and established a dimensionless length scale. For unstable conditions this scale was within 20 percent, but for neutral conditions the error was nearly 45 percent. Because high winds are most frequently associated with a neutral atmosphere, this result would suggest caution in using an exponential approximation for many practical studies.

[Mackey and Ko, 1975] fit their typhoon data which leveled off instead of going to zero with the more complicated expression

$$R(\tau) = a_0 e^{-A\tau} + a_1 \cos(m\tau). \tag{9}$$

They claimed that the addition of the cosine term gave a good representation to their data.

[Cramer, 1959] tried to fit autocorrelations of the u component by the equation

$$1-R(\tau) = c\tau^{2/3} \tag{10}$$

where the constant c is selected to fit the data. Cramer discovered that such a law fit some daytime experiments where the level of turbulence was high. For other cases, Equation (10) was not even approximately valid beyond a few seconds.

[Frost et al., 1978] suggest an even more complicated function for the longitudinal correlation function

$$R(\Delta x) = \frac{2^{2/3}}{\Gamma(1/3)} \left(\frac{\Delta x}{aL_p}\right)^{1/3} K_{1/3} \left(\frac{\Delta x}{aL_p}\right), \tag{11}$$

where K is the modified Bessel function of the second kind and L_p is the longitudinal isotropic turbulence integral scale. Equation (11) is referred to as the von Karman longitudinal correlation function, and a simple exponential model as the Dryden longitudinal correlation function. The Dryden function is more commonly used because there is no compelling evidence that the von Karman model is better, and the Dryden function is much simpler.

IV. THE EKMAN LAYER

Most of the previous discussion has been concerned with the lowest tens of meters of the atmosphere, where the horizontal wind stresses are assumed to be nearly constant and the wind does not turn significantly with height. This layer is sometimes called the surface boundary layer, constant stress layer, or constant flux layer.

The planetary boundary layer, which is also called the friction layer or the atmospheric boundary layer, extends from the surface of the earth to the geostrophic wind level [Huschke, 1959]. The planetary boundary layer includes the surface boundary layer and the Ekman layer. Above the geostrophic wind level is the free atmosphere.

The Ekman layer lies between the surface boundary layer and the free atmosphere. An idealized mathematical description of the wind distribution in this layer is called the Ekman spiral [Huschke, 1959]. This Ekman spiral is

derived by assuming that within the planetary boundary layer the eddy viscosity, K, and density, ρ, are constant. The motion is assumed to be horizontal and steady, the isobars are straight and parallel, and the geostrophic wind is constant with height. The geostrophic wind is represented by the equation

$$U_g = -\frac{1}{\rho f}\frac{\partial p}{\partial n} \tag{12}$$

where U_g is the speed of the geostrophic wind, f is the Coriolis parameter, ρ is pressure, and n is horizontal distance perpendicular to the flow. The n axis increases to the left of the flow in the northern hemisphere. If the x direction is now taken as parallel to the isobars and positive in the direction of the geostrophic wind, one can derive the equations [Hess, 1959]

$$u = U_g\,(1-e^{-az}\cos az) \tag{13a}$$

and

$$v = U_g\,e^{-az}\sin az \tag{13b}$$

where $a = \sqrt{f/2K}$. At $z = 0$ the wind speed is zero. The limiting value of the angle of the wind with the isobars as the surface is approached from above is 45 degrees, and the wind points toward lower pressures. The wind vector turns clockwise with altitude in the Northern Hemisphere and becomes parallel to the isobars at the geostropic wind level. At this level, which is near 1 km, the wind speed is slightly greater than the geostropic value.

Another approach is sometimes used by investigators who are interested in levels above the lowest 20 to 40m if they do not wish to go much above 150m. [Panofsky, 1973] shows that

$$u_* = u_{*0} - 6fz \tag{14}$$

to a very good approximation. The symbol u_{*0} represents the surface friction velocity. Under neutral conditions and homogeneous terrain the following can be written:

$$\overline{u} = \frac{u_{*0}}{k}\,\ln\left(\frac{z}{z_0}\right) + 144fz\ . \tag{15}$$

Because high winds are usually associated with neutral stability, Equation (15) may be quite useful for some investigations.

An empirical power law is also frequently used to represent low level winds [Hess, 1959]. This may be written

$$\overline{u} = \overline{u}_1\left(\frac{z}{z_1}\right)^m \tag{16}$$

where $\bar{u}_1$ is the mean wind speed at a reference level z_1. The exponent m has been found empirically to decrease with increasing lapse rate. [Zhang, 1981] compared the power law with the simple logarithmic wind law which applies to neutral conditions. Wind data from a 164-m tower were examined for one year in Nanjing, China. For the height range from 16m to 164m, the power law represented the actual wind speed distribution better than the logarithmic law.

V. SUMMARY

Wind variation with height in the surface boundary layer can be approximated by a logarithmic wind law. This law is particularly useful in many applications because it is quite good when atmospheric stability is neutral, and high wind speeds are typically associated with neutral atmospheric stability. When stability is not neutral, better accuracy can be obtained by using an equation which contains a small stability term in addition to the logarithmic term. If measurements are inadequate to compute the stability, one can estimate it by Pasquill's method which depends upon time of day, cloud cover, and mean wind speed.

The intensity of turbulence, $\sigma_u/\bar{u}$, varies in space and time. It is usually greater over land than over water, and the intensity is greater over rough land surfaces than over smooth terrain. Intensity of turbulences typically decreases rather rapidly in the lowest 20m and decreases slowly with altitude above this level. Intensity of turbulence normally is greater under unstable conditions than under stable conditions.

Some investigators prefer to measure intensity of turbulence by $\sigma_v/\bar{u}$ instead of $\sigma_u/\bar{u}$. Near the surface σ_v/σ_u is less than unity in slightly unstable, neutral, and stable conditions. In moderately unstable conditions the ratio is near unity. As the atmosphere becomes very unstable, σ_v/σ_u becomes greater than one. At higher altitudes σ_v/σ_u is typically near unity.

The autocorrelation function of u is often irregular in individual cases, but is somewhat smoother when a mean over a large amount of data is taken. A simple exponential function is sometimes used to approximate the decay of the autocorrelation with time or distance, but there is evidence that this can lead to errors of 20 to 45 percent.

Above the surface boundary layer is the Ekman layer where the wind approximately follows an Ekman spiral. The planetary boundary layer consists of this Ekman layer and the surface boundary layer. In the free atmosphere above the planetary boundary layer surface friction with the earth has a negligible influence.

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